Vapor of vapors, saith Koheleth;
vapor of vapors, all is vapor.
—Ecclesiastes 1:2

18. EVAPORATION FROM BARE SOIL AND WIND EROSION

INTRODUCTION

Evaporation in the field can take place from plant canopies, from the soil surface, or from a free-water surface. Evaporation from plants, commonly called *transpiration,* is the principal mechanism of soil-water transfer to the atmosphere when the soil surface is covered with vegetation (see Chapter 19). When the surface is at least partly bare, evaporation can take place from the soil directly. Because it is generally difficult to separate these two interdependent processes, they are often lumped together and treated as if they were a single process, called *evapotranspiration.* Some scientists, however, object to the latter term, believing it to be both cumbersome and unnecessary; they refer to all processes of vapor transfer to the atmosphere—from soil and plants alike—as evaporation (Monteith, 1973).

In the absence of vegetation, and when the soil surface is subject to radiation and wind effects, evaporation occurs entirely from the soil. It is a process that, if uncontrolled, can involve very considerable losses of water in both irrigated and unirrigated agriculture. Under traditionally managed annual field crops, the soil surface often remains largely bare throughout the periods of tillage, planting, germination, and early seedling growth, periods in which evaporation can deplete the moisture of the surface soil and thus hamper the growth of young plants during their most vulnerable stage. Rapid drying of a seedbed can thwart germination and thus doom an entire crop from the outset. The problem can also be acute in young orchards, where the soil surface is often kept bare continuously for several years, as well as in dryland farming in arid zones, where the land is regularly fallowed for several months to collect and conserve rainwater from one season to the next. Finally, direct evaporation from the soil surface is also important in natural ecosystems, wherever the the soil is sparsely vegetated or cleared by animals and by human encroachment.

Evaporation of soil moisture involves not only loss of water but also the danger of soil salination. This danger is greatest in arid areas where annual rainfall is low, irrigation water is brackish, and the groundwater table is high.

PHYSICAL CONDITIONS

Three conditions are necessary for evaporation to occur and persist (Jones, 1991). First, there must be a continual supply of heat to meet the latent heat requirement (about 2.5×10^6 J/kg, or 590 cal/g of water evaporated at 15°C). This heat can come from the body itself, thus causing it to cool, or—as is more commonly the case—it can come from the outside in the form of radiated or advected energy. Second, the vapor pressure in the atmosphere over the evaporating body must remain lower than the vapor pressure at the surface of that body (i.e., there must be a vapor-pressure gradient between the body and the atmosphere), and the vapor must be transported away by diffusion or convection, or both. These two conditions—namely, supply of energy and removal of vapor—are generally external to the evaporating body and are influenced by meteorological factors such as air temperature, humidity, wind velocity, and radiation, which together determine the *atmospheric evaporativity* (the maximal flux at which the atmosphere can vaporize water from a free-water surface).[1]

The third condition for evaporation to be sustained is that there be a continual supply of water from or through the interior of the body to the site of evaporation. This condition depends on the content and potential of water in the body as well as on its conductive properties, which together determine the maximal rate at which the body can transmit water to the evaporation site (usually, the surface). Accordingly, the actual evaporation rate is determined either by the *evaporativity* of the atmosphere or by the soil's own ability to deliver water (sometimes called the *evaporability* of soil moisture), whichever is the lesser (and hence the rate-limiting factor).

If the top layer of soil is initially quite wet, as it typically is at the end of an infiltration episode, the process of evaporation will generally reduce soil wetness and thus increase matric suction at the surface. This, in turn, will cause soil moisture to be drawn upward from the layers below, provided they are sufficiently moist.

1. Atmospheric evaporativity, also called *the evaporative demand of the atmosphere,* is not entirely independent of the properties of the body from which evaporation takes place. For instance, the net supply of energy for evaporation is affected by the slope, surface roughness, reflectivity, emissivity, and thermal conductivity of the soil. Hence the rate of evaporation from a wet soil will not be exactly equal to evaporation from a free-water surface. The latter itself depends on the size, depth, and turbidity of the water body considered. For these reasons, the idea that "atmospheric evaporativity" or "potential evaporation" exists as an exclusively climatic or meteorological "forcing" is only an approximation.

Among the many sets of circumstances under which evaporation may take place are the following:

1. A shallow groundwater table may be present at a constant or variable depth—or it may be absent (or too deep to affect evaporation). Where a groundwater table occurs close to the surface, continual flow may take place from the saturated zone beneath through the unsaturated soil to the surface. If this flow is more or less steady, continued evaporation can occur without materially changing the soil-moisture content (though cumulative salination may take place at the surface). In the absence of shallow groundwater, on the other hand, the loss of water at the surface and the resulting upward flow of water in the profile will necessarily be a transient-state process tending gradually to dry out the soil.
2. The soil profile may be uniform (homogeneous and isotropic). Alternatively, soil properties may change gradually in various directions, or the profile may consist of distinct layers differing in texture or structure.
3. The profile may be shallow, resting on bedrock or some other impervious floor, or it may be deep. If the bottom of the profile is so deep as to remain unaffected by conditions or processes at the surface, the profile is called *semi-infinite*. In intermediate cases, the soil profile may be effectively semi-infinite for a time, then become finite as the downward-propagating effect of the evaporation process reaches the bottom boundary.
4. The flow pattern may be one-dimensional (vertical), or it may be two- or three-dimensional, as in the presence of vertical cracks that form secondary evaporation planes inside the profile.
5. Conditions may be nearly isothermal or strongly nonisothermal. In the latter case, temperature gradients and the conduction of both heat and vapor through the system may interact with liquid water flow.
6. External environmental conditions may remain constant or fluctuate. Such fluctuation, furthermore, can be predictably periodic and regular (e.g., diurnal or seasonal) or it can be highly irregular (e.g., spells of cool or warm weather varying in timing, duration, and intensity).
7. Soil-moisture flow may be governed by evaporation alone or by both evaporation at the top of the profile and internal drainage (or redistribution) at the bottom of the profile.
8. The soil may be stable or unstable. For instance, the surface zone may become denser under traffic or under raindrop impact and subsequent shrinkage. Additionally, the soil surface may become encrusted or infused with salt, which then precipitates as the soil solution evaporates. This is quite apart from the fact that, as a soil dries, its thermal properties inevitably change, including its thermal conductivity and reflectivity.
9. The surface may or may not be covered by a layer of mulch (e.g., plant residues) differing from the soil in hydraulic, thermal, and diffusive properties.
10. Finally, the evaporation process may be continuous over a prolonged period of time or it may be interrupted by regularly recurrent or sporadic episodes of rewetting (e.g., intermittent rainfall or scheduled irrigation).

To be studied systematically, each of these circumstances, as well as others not listed but perhaps equally relevant, must be formulated in terms of a specific set of initial and boundary conditions. A proper formulation of an evaporation process should account for spatial and temporal variability, as well as for interactions with the aboveground and belowground environment. We now proceed to describe in detail a few of the circumstances under which evaporation of soil moisture may occur. We begin with a description of capillary rise, which is often a precursor and contributor to the process of evaporation especially in the presence of a high water table.

CAPILLARY RISE FROM A WATER TABLE

The rise of water in the soil from a free-water surface (i.e., a water table) has been termed *capillary rise*. This term derives from the *capillary model*, which regards the soil as analogous to a bundle of capillary tubes, predominantly wide in the case of a sandy soil and narrow in the case of a clay soil. Accordingly, the equation relating the equilibrium height of capillary rise h_c to the radii of the pores is

$$h_c = (2\gamma \cos\alpha)/r\rho_w g \qquad (18.1)$$

where γ is the surface tension, r the capillary radius, ρ_w the water density, g the gravitational acceleration, and α the wetting angle, normally (though not always justifiably) taken as zero. This equation predicts that water will rise higher, though typically less rapidly, in a clay than in a sand. The reason why is that the former soil type contains narrower pores. However, soil pores are not individual capillary tubes of uniform or constant radius, and hence the height of capillary rise will differ in different pores. Above the water table, matric suction will generally increase with height. Consequently, the number of water-filled pores, and hence wetness, will diminish in each soil as a function of height. The rate of capillary rise, that is, the flux, generally decreases with time as the soil is wetted to greater height and as equilibrium is approached.

The wetting of an initially uniformly dry soil by upward capillary flow, illustrated in Fig. 18.1, is a rare occurrence in the field. In its initial stages, this process is similar to infiltration, except that it takes place in the opposite direction, that is to say, against the direction of gravity. At later stages of the process, the flux does not tend to a constant value, as in downward infiltration, but to zero. The reason why is that the direction of the gravitational gradient is opposite to the direction of the matric suction gradient, and when the latter (which is large at first but decreases with time) approaches the magnitude of the former, the overall hydraulic gradient approaches zero.

Such an ideal state of static equilibrium between the gravitational head and the suction head is the exception rather than the rule under field conditions. In general, the condition of soil water is not static but dynamic—constantly in a state of flux rather than at rest. Where a water table is present, soil water generally does not attain equilibrium even in the absence of vegetation, since the soil surface is subject to solar radiation and the evaporative demand of the ambient atmosphere. However, if soil and external conditions are constant,

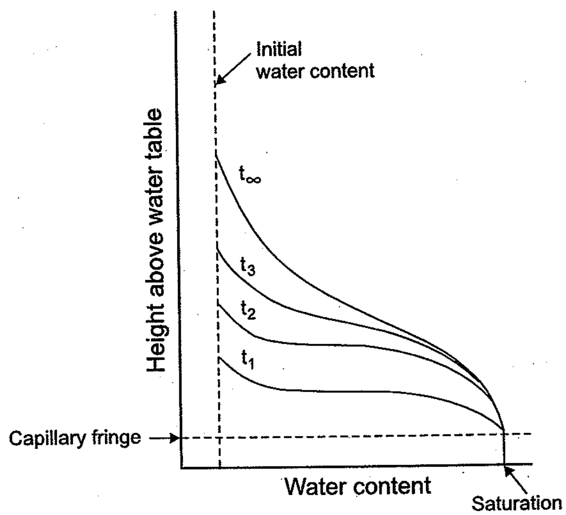

Fig. 18.1. The upward infiltration of water from a water table into a dry soil: water content distribution curves (moisture profiles) for various times ($t_1 < t_2 < t_3 < t_\infty$), where t_∞ is the profile after an infinitely long time (equilibrium). Note that the equilibrium curve is in effect the wetting branch of the soil-moisture characteristic. Note also that what is at first a sharp wetting front, representing the limit of the upward advancing water, gradually becomes diffuse and ends up as a smooth curve characteristic of the particular soil's pore-size distribution.

that is, if the soil is of stable structure, the water table is stationary, and atmospheric evaporativity also remains constant (at least approximately)—then, in time, a steady-state flow situation can develop from water table to atmosphere via the soil. To be sure, we must hasten to qualify this statement by noting that in the field the flow regime will at best be a quasi-steady-state flow, since diurnal fluctuations and other perturbations prevent attainment of truly stable flow conditions. Nevertheless, the representation of this process as a steady-state flow can be a useful approximation from the analytical point of view.

STEADY EVAPORATION FROM A SHALLOW WATER TABLE

The steady-state upward flow of water from a water table through the soil profile to an evaporation zone at the soil surface was first studied by Moore (1939). Solutions of the flow equation for this process were given by several workers, including Philip (1957d), Gardner (1958), Anat *et al.* (1965), and Ripple *et al.* (1972), and Hillel (1977).

The equation describing steady upward flow is

$$q = K(\psi)\,(d\psi/dz - 1) \quad \text{or} \tag{18.2}$$

$$q = D(\theta)\,d\theta/dz - K(\psi) \tag{18.3}$$

where q is flux (equal to the evaporation rate under steady-state conditions), ψ suction head, K hydraulic conductivity, D hydraulic diffusivity, θ volumetric wetness, and z height above the water table. The equation shows that the flow stops ($q = 0$) when the suction profile is at equilibrium ($d\psi/dz = 1$). Another form of Eq. (18.2) is

$$q/K(\psi) + 1 = d\psi/dz \tag{18.4}$$

Integration should give the relation between depth and suction or wetness:

$$z = \int \frac{d\psi}{1 + q/K(\psi)} = \int \frac{K\psi}{K(\psi) + q} d\psi \quad \text{or} \tag{18.5}$$

$$z = \int \frac{D(\theta)}{K(\theta) + q} d\theta \tag{18.6}$$

To perform the integration in Eq. (18.5), we must know the functional relation between K and ψ, that is, $K(\psi)$. Similarly, the functions $D(\theta)$ and $K(\theta)$ must be known if Eq. (18.6) is to be integrated. An empirical equation for $K(\psi)$, given by Gardner (1958), is

$$K(\psi) = a(\psi^n + b)^{-1} \tag{18.7}$$

where parameters a, b, n are constants that must be determined for each soil. In this formulation, the suction head ψ is expressed in terms of centimeters. Accordingly, Eq. (18.2) becomes

$$e = q = a(d\psi/dz - 1)/(\psi^n + b) \tag{18.8}$$

where e is the evaporation rate.

With Eq. (18.7), Eq. (18.5) can be used to obtain suction distributions with height for different fluxes, as well as fluxes for different surface-suction values. The theoretical solution is shown graphically in Fig. 18.2 for a fine sandy loam soil with an n value of 3. The curves show that the steady rate of capillary rise and evaporation depends on the depth of the water table and on the suction at the soil surface. This suction is dictated largely by the external conditions: The greater the atmospheric evaporativity, the greater the suction at the soil surface on which the atmosphere is acting. However, increasing the suction at the soil surface, even to the extent of making it infinite, can increase the flux through the soil only up to an asymptotic maximal rate, which depends on the depth of the water table. Even the driest and most evaporative atmosphere cannot steadily extract water from the surface any faster than the soil profile can transmit from the water table to that surface. The fact that the soil profile can limit the rate of evaporation is a remarkable and useful feature of the unsaturated flow system. The maximal transmitting ability of the profile depends on the hydraulic conductivity of the soil in relation to the suction, $K(\psi)$.

Disregarding constant b of Eq. (17.7), Gardner (1958) obtained the function

$$q_{\max} = Aa/d^n \tag{18.9}$$

where d is the depth of the water table below the soil surface, a and n are constants from Eq. (18.7), A is a constant that depends on n, and $q_{\max}$ is the limiting (maximal) rate at which the soil can transmit water from the water table to the evaporation zone at the surface.

We can now see how the actual steady evaporation rate is determined either by the external evaporativity or by the water-transmitting properties of the soil, depending on which of the two is lower and therefore limiting. Where the water table is near the surface, the suction at the soil surface is low and the evaporation rate is dictated by external conditions. However, as the water table becomes deeper and as the suction at the soil surface increases, the evaporation rate approaches a limiting value regardless of how high external evaporativity may be.

Equation (18.9) suggests that the maximal evaporation rate decreases with water-table depth more steeply in coarse-textured soils (in which n is greater because the conductivity falls off more steeply with increasing suction) than in clayey soils. Nevertheless, a sandy loam soil can still evaporate water at an appreciable rate (as shown in Fig. 18.2) even when the water table is as deep as 1.80 m. Figure 18.3 illustrates the effect of texture on the limiting evaporation rate.

The subsequent findings of numerous workers (e.g., Talsma, 1963) have generally accorded with the theory just discussed. Hadas and Hillel (1968), however, found that experimental soil columns deviated from predicted behavior, apparently owing to spontaneous changes of soil properties (e.g., structure, albedo, salinity), particularly at the surface, during the course of evaporation.

Anat *et al.* (1965) developed a modified set of equations employing dimensionless variables. [For basic explanations of the principles and criteria involved in the dimensionless approach, see Corey (1977) and Miller (1980).]

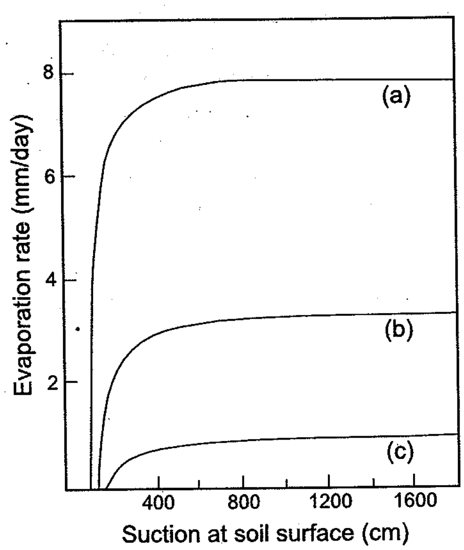

Fig. 18.2. Steady upward flow and evaporation from a water-table as a function of the suction prevailing at the soil surface, with water-table at: (a) 90-cm depth, (b) 120-cm, (c) 180-cm. The soil is a fine sandy loam, with $n = 3$. (After Gardner, 1958.)

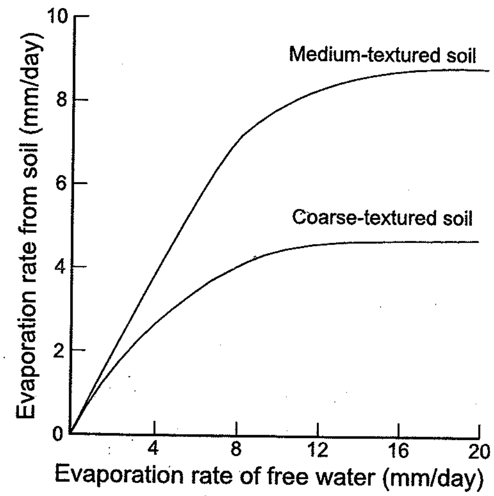

Fig. 18.3. Theoretical relation between evaporation rate from coarse- and medium-textured soils (water-table depth-60 cm) and evaporation rate from free-water surface. (After Gardner, 1958.)

Their theory also leads to a maximal evaporation rate e_{max} varying inversely with water-table depth d to the power of n:

$$e_{max} = [1 + 1.886/(n^2 + 1)]\, d^{-n} \tag{18.10}$$

An early theoretical analysis of steady evaporation from a two-layered soil profile was carried out by Willis (1960), who used a graphical method of solution. He made the following assumptions: (a) The steady flow through the layered profile is governed only by the transmission properties of the profile (external evaporativity taken to be infinite); (b) matric suction is continuous at and through the interlayer boundary, though wetness and conductivity may be discontinuous (i.e., change abruptly); (c) the same empirical $K(\psi)$ function given by Eq. (18.7) holds for both layers, but the values of parameters a, b, and n are different; and (d) each soil layer is internally homogenous. With these assumptions, Eq. (18.2) leads to

$$\int_0^L dz + \int_L^{d+L} dz = \int_{\psi 0}^{\psi L} \frac{d\psi}{1 + e/K_1(\psi)} + \int_{\psi L}^{\psi L+d} \frac{d\psi}{1 + e/K_1(\psi)} \tag{18.11}$$

where L and d are the thicknesses of the bottom and top layers, respectively. The integral in this equation relates water table depth $(L + d)$ to the suction at the soil surface for any given evaporation rate. Assuming that suction at the surface is infinite, one can calculate the limiting (maximal) evaporation rate for any given water-table depth and profile-layering sequence.

All of the preceding treatments apply to cases in which soil properties alone determine the evaporation rate. A more realistic approach must include the role of meteorological conditions. A flexible treatment of steady-state evaporation, based on numerical (rather than analytical or graphical) methods of solution, was developed by Ripple *et al.* (1972). Their results are illustrated in

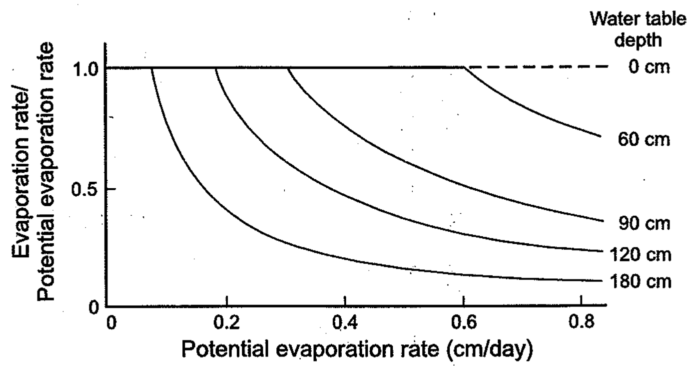

Fig. 18.4. Dependence of relative evaporation rates e/e_{pot} on the potential evaporation rate for a clay soil. Numbers at curves indicate depth to water table in centimeters. (After Ripple *et al.*, 1972.)

Figs. 18.4 and 18.5. Their procedure makes it possible to estimate the steady-state evaporation from bare soils (including layered ones) with a high water table. The required data include soil-moisture characteristic curves, water-table depth, and standard records of air temperature, air humidity, and wind velocity. The theory takes into account both the relevant atmospheric factors, and the soil's capability to transmit water in liquid and vapor forms. The possible effects of thermally induced water transfer (except in the vapor phase) and of salt accumulation at the surface are additional factors to be taken into account. Additional solutions for steady-state evaporation from a shallow water table were given by Warrick (1988).

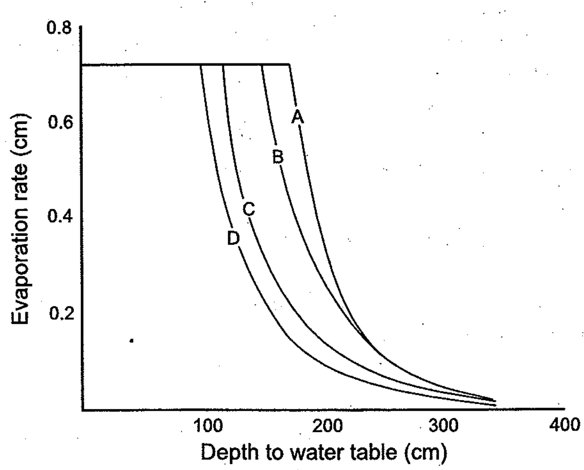

Fig. 18.5. Influence of layering on the relation between evaporation rate and depth to water table. Limiting curves of soil-water evaporation are shown for a homogeneous soil (A), a two-layer soil with upper-layer thickness of either 3 cm (B) or 10 cm (C), and a three-layer soil with the thickness of the intermediate and uppermost layers equal to 10 cm each (D). (After Ripple *et al.*, 1972.)